

1 Modeling the thickness of perennial ice covers on stratified lakes of the Taylor Valley,
2 Antarctica

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14 **Abstract**

15 A ~~one-dimensional~~ ice cover model was developed to predict and constrain drivers of long-
16 term ice thickness trends in chemically stratified lakes of Taylor Valley, Antarctica. The model is
17 driven by surface radiative heat fluxes and heat fluxes from the underlying water column. The
18 model successfully reproduced 16 ~~years~~ (between 1996 and 2012) of ice thickness changes for
19 west lobe of Lake Bonney (average ice thickness = 3.53 m; RMSE = 0.09 m, n = 118) and Lake
20 Fryxell (average ice thickness = 4.22 m; RMSE = 0.21 m, n = 128). Long-term ice thickness
21 trends require coupling with the thermal structure of the water column. The heat stored within
22 the temperature maximum of lakes exceeding a liquid water column depth of 20 m can either
23 impede or facilitate ice thickness change depending on the predominant climatic trend
24 (~~temperature~~ cooling or warming). As such, shallow (< 20 m deep water columns) perennially
25 ice-covered lakes without deep temperature maxima are more sensitive indicators of climate
26 change. The long-term ice thickness trends are a result of surface energy flux and heat flux from
27 the deep temperature maximum in the water column, the latter of which results from absorbed
28 solar radiation.
29

30

31 **Introduction**

32 Perennially ice-covered lakes in the McMurdo Dry Valleys, Antarctica have long been
33 studied as novel habitats for life. Ice covers offer protection from wind-driven turbulence, and
34 together with strong salinity gradients within the lakes, permit permanent vertical temperature
35 stratification, the development of supersaturated gas concentrations (Priscu and others, 1996;
36 Andersen and others, 1998; Wharton and others, 1986), and altered sediment pathways (Doran
37 and others, 1994). Although the permanent ice covers provide insulation from the atmosphere,
38 they do allow penetration of solar radiation, often producing thermal maxima deep in the water
39 column in association with saltier water (Vincent and others, 2008). Despite the deep
40 temperature maxima, the salt gradient maintains vertically stable water columns (Spigel &
41 Priscu, 1996). Ice cover thickness is balanced primarily by ablation at the surface and ice growth
42 at the bottom (Chinn, 1993; Adams and others, 1998). McKay and others (1985) developed a
43 steady state energy-balance model that predicts ice cover thickness based on surface ablation
44 (mass loss by all means) rates and latent heat of ice formation. Latent heat release, associated
45 with ice growth, is conducted upwards and lost to the atmosphere due to ablation (McKay and
46 others, 1985). Because of the tight coupling of lake ice thickness to atmospheric processes, the
47 ice covers should respond directly to changing climate (Launiainen & Cheng, 1998; Reid &
48 Crout, 2008; Wharton and others, 1992).

49 Lakes in Taylor Valley, one of the McMurdo Dry Valleys have been perennially ice-covered
50 since at least 1903, when they were discovered by Robert Falcon Scott (Scott, 1905). Taylor
51 Valley lake volumes are controlled by a balance of summer temperatures warm enough to
52 generate melt from local glaciers, yet cold enough for the lake ice covers to persist through the
53 summer (Chinn, 1993). The ice thickness in Taylor Valley lakes has been fluctuating between 3
54 and 6 m over the past two decades (McKay and others, 1985; Doran and others, 2002b). Ice
55 thickness of Lake Hoare (freshwater lake) in McMurdo Dry Valleys thinned from 1977 to 1986
56 at a rate of $>0.28 \text{ m a}^{-1}$ (Wharton and others, 1989); however little climate data is available for
57 this time period. Between 1986 and 1999, its lake-ice thickness increased at a rate of 0.11 m a^{-1}
58 in association with an annual air temperature decrease of $0.7^\circ\text{C per decade}$ (Doran and others,
59 2002b). This period of pronounced cooling was terminated by one of the warmest summers on
60 record, resulting in a significant lake level rise (Doran and others, 2008; Barrett and others,
61 2008). Following this warm pulse, lake-ice thicknesses have been decreasing since early- to mid-
62 2000 at rates between 0.07 and 0.18 m a^{-1} (unpublished data). Here we present results from a
63 lake ice model that allowed us to differentiate between the influence of atmospheric processes
64 (i.e., changes in climate) and the long-term heat storage in the water column on lake ice
65 thickness.

66

67 **Site Description**

68 The McMurdo Dry Valleys, East Antarctica is an ice-free region with perennially ice-
69 covered lakes typically located in hydrologically closed-basins. This arid environment is a result
70 of combination of low precipitation and the Transantarctic Mountains blocking the ice sheet flow
71 (Chinn, 1993). The mean annual valley-bottom temperatures range from -14.8°C to -30.0°C ,
72 with summer temperatures high enough to enable glacial melt, which feeds into the lakes (Doran
73 and others, 2002a). The two lakes in this study, west lobe of Lake Bonney (WLB) and Lake
74 Fryxell (Figure 1), are in different microclimatic zones in Taylor Valley. WLB is located
75 approximately 27 km from the coast and influenced by strong, dry winds from the Antarctic

76 plateau, whereas Lake Fryxell, adjacent to the coast, is influenced by moist coastal winds
77 (Fountain and others, 1999; Doran and others, 2002a). Lake levels in Taylor Valley have been
78 rapidly increasing since early 2000's and the depths of WLB and Lake Fryxell, as of Jan 2012
79 are 41.2 and 22.4 m ~~deep~~, respectively (data obtained from www.mcmlter.org). The average ice
80 thickness between 1996 and 2012 of WLB and Lake Fryxell is 3.5 and 4.2 m, respectively.

81

82 **Model Development**

83 *General Model Description*

84 Launiainen and Cheng (1998) provide a general description and evolution of different ice
85 models since thermodynamic modeling of ice began in the late 1800's. Despite extensive
86 research, no universal lake ice model is available, as each lake has different characteristics and
87 unique adjustments are necessary (Reid & Crout, 2008). The model we developed uses
88 generalized parameters to make it adaptable to other stratified perennially ice-covered lakes.

89 The model developed here is modified from an approach by Launiainen and Cheng (1998)
90 and Reid and Crout (2008). The model considers 1) the dynamic coupling of atmospheric
91 processes with ice (surface energy balance), 2) the dynamic coupling of the water column with
92 ice (moving boundary condition), and 3) heat conduction within the ice and the water column
93 (two ~~one-dimensional~~ heat equations) (Figure 2). Three boundaries are defined in the model:
94 atmosphere/ice, ice/water, and water/bedrock. The vertical temperature profile of the ice and the
95 water column is solved using a spectral method based on the state of the boundaries. The main
96 advantage of the spectral method over the finite difference or the finite element method is a
97 higher degree of accuracy for smooth problems allowing a coarse computational grid (Trefethen,
98 2000).

99 Surface ice temperature was calculated based on the summation of surface radiative fluxes
100 and becomes the top boundary condition for the ~~one-dimensional~~ heat equation (~~see Surface~~
101 *Energy Balance* section for details). Known surface ice temperature allows for the calculation of
102 temperature profiles of the ice as the temperature at the bottom of the ice is set to the freezing
103 point. The water temperature profile is solved using a second ~~one-dimensional~~ heat equation,
104 with penetrating solar radiation acting as the heat flux contribution to the water column
105 (water/bedrock boundary temperature is fixed). Ice and water temperature profiles are coupled at
106 the ice/water interface accounting for the conductive, latent, and sensible heat fluxes. The
107 ice/water interface is a moving Dirichlet boundary condition, which forces the solution of a heat
108 equation to converge at the freezing point of the water. This boundary incorporates latent heat
109 fluxes associated with the phase transition (Gupta, 2003).

110 A ~~one-dimensional~~ solution to the heat equation is valid because it succinctly approximates
111 the physical processes. The lateral thermal gradient can be considered negligible with regards to
112 depth, allowing a ~~one-dimensional~~ heat equation to be employed because vertical exchanges are
113 much greater than lateral exchanges. Heat exchange within the ice is assumed to occur through
114 conduction (even though this assumption can break down when ice is isothermal) and is based on
115 a temperature gradient defined at the atmosphere/ice and ice/water boundaries. The heat within
116 the ice diffuses vertically at a rate defined by the thermal diffusivity of the ice and the
117 temperature gradient. Similar to the approach used for the ice, a ~~one-dimensional~~ heat equation is
118 used to calculate the temperature profile of the water column in order to better constrain heat
119 fluxes at the ice/water boundary. The water column of Taylor Valley lakes we modeled are
120 chemically stratified and vertical diffusive heat transfer occurs at the molecular level; our study
121 lakes have no vertical convective mixing (Spigel & Priscu, 1998). Horizontal motion in the lakes

122 can occur as the result of horizontal advection caused by lateral stirring induced by contact with
123 glacial ice (WLB) and stream input during the austral summer (Spigel & Priscu, 1998) and are
124 not considered in this model.

125 The model employs dimensional analysis permitting the grouping of variables that describe
126 physical parameters in a dimensionless form. Dimensional analysis is based on the Buckingham
127 π -theorem, allowing for the reduction of independent variables by k dimensions, where k
128 represents physical dimensions required to describe n variables used in the model. Then, it
129 follows that there will be π dimensionless groups based on $n - k$. As a result, this method reduces
130 the complexity of the model and, consequently, reduces the complexity of the solution to the heat
131 equation and allows the computational domain to be fixed at [-1, 1], regardless of the ice
132 thickness.

133 134 *Model input data*

135 Meteorological data used in the model include air temperature, shortwave radiation (400 to
136 1100 nm), relative humidity, and wind speed. Longwave upwelling and downwelling radiation,
137 and turbulent sensible and latent heat fluxes were parameterized using equations adapted from
138 Reid and Crout (2008), discussed in more detail below. The albedo of the ice was simulated as a
139 spline function based on a five-month data record (September to February) from WLB, obtained
140 from Fritsen and Priscu (1999). Based on this record, the largest albedo change (increase) occurs
141 during the summer months when ice morphology and hoar frost in air bubbles change due to the
142 thermal properties of the ice (Fritsen & Priscu, 1999; Adams and others, 1998). The missing
143 albedo data for the remainder of the year (spring and fall) were extrapolated. Any error in this
144 assumption is reduced by the diminished importance of radiation during these periods and the
145 complete lack of radiation during the winter months.

146 The model was run with 16 years of meteorological data obtained from Lake Bonney's
147 meteorological weather station, located on the southern shore of east lobe of Lake Bonney
148 (Figure 1). Data were collected using Campbell Scientific dataloggers (CR10 and 21X) at 30 s
149 sampling intervals and stored as 15 min averages (Doran and others, 1995; Doran and others,
150 2002a). Air temperature was measured using Campbell 207 and 107 temperature probes at a 3 m
151 height (Doran and others, 1995). Shortwave radiation was obtained with Li-Cor LI200S and
152 LI200X silicon pyranometer (Doran and others, 1995). Wind speed was obtained with R.-M.
153 Young model 05103 wind monitors (Doran and others, 1995). The model was run at 12-hour
154 intervals, dynamically averaging high-resolution data (15 min intervals) during simulations.

155 Gaps in data occurred, ranging from hours to months, due to either failed sensors or failed
156 dataloggers. Missing shortwave radiation was remedied by using well-known published
157 equations (see *Shortwave Radiation* section) dynamically substituting gaps in data during model
158 simulations. Missing air temperature values were substituted by using a function based on daily
159 averaged temperatures from 16 years of data. Missing relative humidity and wind speed data
160 represent cumulative 6.6 and 4.6 % of the entire dataset, respectively, and were determined by a
161 uniform random number generator (within the upper and lower limit of each variable). Uniform
162 distribution ensures that a value within the set range has equal probability of being generated.

163 164 *Validation data*

165 Ice thickness data were measured manually at the drill hole sites from the bottom of the ice to
166 the piezometric water level, representing the ice thickness ~~in water equivalents~~. Due to logistical
167 constraints, the ice thickness data were only available during the austral summers when the ice

168 was most dynamic, with the exception of an extended season in 2008, when data were obtained
 169 through late fall (into March). Replicate ice thickness measurements (3 to 10) were made during
 170 the same day (within a 10 m² grid) or within two days of each other, and show a large variability
 171 (largest standard deviation of seasonally measured ice thickness is 0.54), introducing error during
 172 the validation procedure of the model.

173 Continuous ablation data were measured from suspended pressure transducers at the bottom
 174 of the ice (Dugan and others, 2013). Data were recorded at one-minute intervals (averaged over
 175 20 min), using Druck PDCR 1830 or Keller Series 173 pressure transducer and stored on
 176 Campbell Scientific CR10X and CR1000 dataloggers (Dugan and others, 2013).

177 *Surface Energy Balance*

178 The vertical temperature profile of the ice and its evolution over time can be modeled
 179 utilizing a one-dimensional unsteady heat equation:

$$182 \quad pc \frac{\partial T(z,t)}{\partial t} = - \frac{\partial}{\partial z} \left(-k \frac{\partial T(z,t)}{\partial z} + I(z,t) \right) \quad (1)$$

183 where p is the density of ice or water, c is the specific heat capacity of ice or water, T is
 184 temperature, z is depth, t is time, k is thermal conductivity, and I is the heat source term within
 185 the ice and the water column. All constants and variables are summarized in Appendix A.
 186 Shortwave radiation is the only significant flux that contributes to the I term because ice is
 187 opaque in infrared and transparent in the visible spectrum (McKay and others, 1994). The
 188 transmittance of sunlight through the perennially ice-covered lakes in Taylor Valley can be
 189 approximated using the Lambert-Beer law (McKay and others, 1985), by defining a constant
 190 extinction coefficient. However, this technique does not account for the absorbed near-infrared
 191 radiation at the very surface of the ice (Bintanja & Vandenbroeke, 1995; Brandt & Warren,
 192 1993). A method used in Bintanja and Vandenbroeke (1995) was adapted that parameterized
 193 solar energy by partitioning absorbed solar energy at the very surface and subsurface of the ice,
 194 expressing the heat source term as:

$$197 \quad I(z,t) = (1 - \chi)(1 - \alpha)Q_s e^{-\kappa z} \quad (2)$$

198 where χ accounts for a fraction of absorbed shortwave radiation at the surface, α is albedo, Q_s is
 199 shortwave radiation, and κ is the extinction coefficient of the ice at depth z .

200 The top boundary condition for Eq. (1) is calculated based on the summation of all surface
 201 radiative heat fluxes:

$$204 \quad F(T_s) = \chi(1 - \alpha)Q_s + Q_b + Q_d + Q_h + Q_l + F_i \quad (3)$$

205 where T_s is ice surface temperature, Q_s is shortwave radiation, Q_b is longwave upwelling
 206 radiation, Q_d is longwave downwelling radiation, Q_h is sensible heat flux, Q_l is latent heat flux,
 207 and F_i is the heat conduction at the ice surface. However, T_s cannot be solved based on this
 208 equation alone because the longwave upwelling radiation, sensible, and conductive heat fluxes
 209 depend on the surface ice temperature. The Newtonian iterative method is implemented in the
 210 model to solve for T_s , which is used to determine the temperature gradient within the ice cover.
 211 Positive fluxes are toward the ice, negative are away from the ice.

212
 213

214 *Shortwave Radiation*

215 Shortwave radiation (Q_s) for an atmosphere-less Earth is defined as:

216
217
$$Q_s = S \cos Z \quad (4)$$

218
219 where S is a solar constant and Z is a solar zenith, derived using solar declination, solar angle,
220 and latitude. Atmospheric attenuation of Q_s requires modification of Eq. (4) with empirical
221 parameters related to the study site and numerous equations are available in the literature (Lumb,
222 1964; Bennett, 1982; Reid & Crout, 2008). Parameterization of Q_s for an Antarctic site has been
223 done by Reid and Crout (2008) and was adopted here.

224
225 *Longwave Radiation*

226 Longwave upwelling radiation (Q_b) was modeled based on the Stefan-Boltzmann law as a
227 function of surface ice temperature and is expressed as:

228
229
$$Q_b = -\varepsilon_i \sigma_s T_s^4 \quad (5)$$

230
231 where ε_i represents dimensionless surface emissivity and σ_s is the Stefan-Boltzmann constant
232 (Launiainen & Cheng, 1998).

233 Downwelling longwave radiation (Q_d) takes a form of Eq. (5) but is controlled by surface air
234 temperature (T_a) and cloud cover (C) (Guest, 1998; Konig-Langlo & Augstein, 1994). Cloud
235 cover is expressed as a unitless value (from 0 to 1) representing the fraction of sky covered by
236 clouds. However, cloud cover data for the WLB is unavailable. Missing cloud cover data can be
237 treated as an adjustable parameter in models driven by surface energy balance (Liston and others,
238 1999). Here, we chose C to be generated by a random number generator (uniform distribution) at
239 daily intervals in order to simplify the model by decreasing the number of adjustable parameters.
240 Downwelling longwave radiation requires parameterization of atmospheric emissivity (ε_a), which
241 accounts for C , vapor distribution, and temperature (Konig-Langlo & Augstein, 1994). The
242 equation for Q_d was adopted from Konig-Langlo and Augstein (1994) because they
243 parameterized ε_a based on Antarctic datasets, which is applicable for high-latitudes:

244
245
$$Q_d = (0.765 + 0.22C^3) \sigma_s T_a^4 \quad (6)$$

246
247 *Sensible and Latent Heat Flux at the Ice Surface*

248 Turbulent sensible (Q_h) and latent ~~heat~~ (Q_l) fluxes at the atmosphere/ice boundary are
249 modeled after Launiainen and Cheng (1998) and are expressed as:

250
251
$$Q_h = \rho_a c_a C_H (\Delta T) V \quad (7)$$

252
253
$$Q_l = \rho_a R_l C_E (\Delta q) V \quad (8)$$

254
255 where ρ_a is air density, c_a is specific heat capacity of air, V is wind speed, R is enthalpy of
256 vaporization (or sublimation if $T_s < 273.15$), ΔT is the temperature difference between the
257 atmosphere and the ice surface, Δq is specific humidity difference between the atmosphere and
258 the ice surface, and C_H and C_E are dimensionless bulk transfer coefficients. Strictly speaking, C_H

259 and C_E depend on atmospheric stability (Launiainen & Cheng, 1998); however, here they are
260 treated as constants approximated based on Parkinson and Washington (1979).

261

262 *Conductive Heat Flux*

263 Conductive heat flux at the surface of the ice is defined as a temperature gradient between the
264 air and the top layer of the ice:

265

$$266 \quad F_i = -k \left(\frac{\partial T}{\partial Z} \right) \quad (9)$$

267

268 where T represents the temperature difference defined by the depth Z , and k is the thermal
269 conductivity of the ice. Similarly, heat conduction between ice and the water takes the form of
270 Eq. (9).

271

272 *Ice ablation and growth*

273 Ice removal from the top of the ice is approximated as a function of net surface radiative heat
274 fluxes, similar to Eq. (3). If the net radiative heat flux (Q_{net}) is greater than zero, the excess
275 energy is used to remove the surface layer of the ice, and can be expressed as:

276

$$277 \quad \frac{d_{ABL}}{dt} = - \left(\frac{1}{Lp} \right) \Sigma Q_{net} \quad (10)$$

278

279 where L is latent heat of freezing and p is ice density. Such an approach does not distinguish
280 between sublimation and ice melt; however, it sufficiently represents cumulative annual ice
281 removal from the surface of the ice (Figure 3).

282

283 Ice growth is only considered at the bottom of the ice, which is controlled by sensible heat
284 flux from the water, conduction of heat at the ice/water boundary, and latent heat flux associated
285 with the phase change (Danard and others, 1984; Launiainen & Cheng, 1998). The temperature
286 at the ice/water interface is assumed to be at the freezing point. Heat fluxes at the ice/water
287 boundary are difficult to determine because sensible and conductive heat fluxes from the water
288 are difficult to ascertain. Sensible heat from the water can be either approximated as a constant if
289 the water temperature beneath the ice is not known (Launiainen & Cheng, 1998) or calculated, if
290 the water temperature beneath the ice is known, using modified Eq. (7) (Reid and Crout, 2008).
291 However, for proper determination of fluxes from the water column, thermal structure of the
292 water profile should be coupled with the ice (Maykut & Untersteiner, 1971; Launiainen &
293 Cheng, 1998). Here, a second heat equation is solved (Eq. (1)) to calculate the water temperature
profile permitting dynamic calculations of heat fluxes between the water and the ice.

294 **Initial Conditions**

295 The initial condition of the ice temperature profile can be approximated as a linear
296 temperature gradient between the ice surface and the ice bottom because the system rapidly
297 adjusts (Reid & Crout, 2008). However, for ice models coupled with ~~the~~ unique water
298 temperature profiles, such as west lobe of Lake Bonney, which takes decades to develop
299 (Vincent and others, 2008), linear approximation might not be appropriate when attempting to
300 model long time series. The initial temperature profiles of the ice and the water column in this
301 model were determined in two steps. First, the temperature gradient between the surface of the
302 ice, ice/water interface, and bottom of the lake was set up as a linear function. The simulation
303 was run until the water temperature profile developed a temperature maximum, similar to that of
304 WLB. Second, the last calculated value, in the previous step, was used as an initial temperature
305 profile for the ice and the water column condition.

306 **Model Validation**

307 The model was validated using 16 years of seasonally measured ice thickness data and 6
308 years of continuous ablation data from WLB. In addition, modeled vs. measured temperature
309 profiles from WLB are qualitatively compared. The versatility of the model was validated using
310 16 years of seasonally measured ice thickness data from nearby Lake Fryxell.

311 The model accurately predicts ice thickness trends for WLB (Figure 4) for the times when ice
312 thickness was measured, including the general trend of increasing ice thickness from 1996 to
313 2002, followed by the subsequent decrease after 2002. The ice thickness during the
314 measurements in March 2008 when the ice was noted to be flooded with lake water, suggests
315 that the seasonal amplitude of ice thickness variations could be poorly resolved by the model
316 under specific conditions. Despite this deviation, measured vs. modeled ice thickness have a
317 correlation coefficient of $r = 0.95$ ($p < 0.001$) and ~~root-mean-square-error (RMSE)~~ of 0.09 m
318 indicating that the model reasonably captures decadal ice thickness trend variability (Figure 5).

319 Despite the simplicity of the ablation calculation, cumulative annual ablation is well
320 predicted (Figure 3) with a correlation coefficient of $r = 0.99$ ($p < 0.001$) and RMSE of 0.11 m
321 between measured and modeled ablation. However, the significance of the fit should be
322 interpreted with caution, as r can be artificially inflated with a large number of data points ($n >$
323 4200). For this reason, RMSE is provided, as it represents an absolute error in units of the data
324 (Willmott, 1981).

325 The model predicted water temperature profile similar to that of WLB. A typical WLB
326 temperature profile (obtained in 2011) is shown in Figure 6A with temperature maximum at
327 approximately 8.5 m depth, which is captured by the modeled temperature maximum (Figure
328 6B).

329 For Lake Fryxell, the model was forced with meteorological data from the Lake Fryxell basin
330 (Figure 1). However, Lake Fryxell ice morphology (and the ice thickness) is significantly
331 different than of WLB and parameters related to optical properties of the ice required adjustment.
332 As such, κ and χ_s were parameterized (increased) during the simulation. Albedo data for Lake
333 Fryxell is not available and values used for WLB simulations were incorporated. Predicted ice
334 thickness for Lake Fryxell shows a good measured vs. modeled correlation of $r = 0.90$ ($p <$
335 0.001) and RMSE of 0.21 m (Figure 7).

336 Model Sensitivity

337 The model sensitivity was tested by increasing/decreasing variables input by 10% of their
338 nominal value, one at a time. The only exception is air temperature, ~~as 10% change would create~~
339 ~~unrealistic conditions. Instead, the air temperature~~ sensitivity was tested by an increase/decrease
340 of ~~temperature by~~ 1°K. A simple modified sensitivity index (S_i) was adapted from Hoffmann and
341 Gardner (1983):
342

$$343 S_i = \left| 1 - \left(\frac{D_{min}}{D_{max}} \right) \right| \quad (11)$$

344 where D_{min} is model output (ice thickness) when a variable was decreased, and D_{max} is the model
345 output when a variable was increased. Sensitivity index values range from 0 to 1, where 1
346 represents high sensitivity to a change in a parameter; values ~~less than~~ 0.01 indicate no
347 sensitivity (Hoffmann & Gardner, 1983). Each test was run over a ~~16-year~~ period or until the ice
348 cover melted, whichever came first. The sensitivity test was performed on shortwave radiation
349 (Q_s), air temperature (T_a), bulk transfer coefficients (C_E and C_H), albedo (α), extinction
350 coefficient (κ), and a parameter responsible for partitioning shortwave radiation at the surface of
351 the ice (χ). Sensitivity indices (S_i) for each parameter are summarized in Table 1, from highest to
352 lowest, and selected parameters are briefly discussed below.
353

354 Shortwave radiation is the largest contributor of the energy at the very surface of the ice. As
355 such, the output of the model is highly sensitive to parameters controlling shortwave radiation,
356 including absorption (χ) and reflectance (α) (Table 1). For example, an increase of χ or α
357 decreases solar radiation that penetrates an ice cover, decreasing the overall energy of the system
358 that otherwise would be available for ice melt. This suggests ~~that~~ the model is highly sensitive to
359 parameters controlling surface optical properties of the ice. Q_s and T_a have an inverse effect on
360 the ice cover thickness. An increase of Q_s and T_a significantly contributes to the surface ice
361 temperature, as well as penetrating solar radiation, both of which will enhance melt.

362 The model is relatively insensitive to changes in bulk transfer coefficients (C_E and C_H ,
363 respectively). Changes in C_E create small variations in the model's output (yet preserving the
364 long-term trends), whereas, changes in C_H make virtually no difference to the output (Table 1).
365 Similar sensitivity results were found by Vincent and others (2008). The sensitivity test validates
366 our original assumption of constant bulk transfer coefficients, reducing the computational
367 complexity of the model.

368 Lake ice thickness is controlled by the surface energy balance and is presumed to reflect
369 climatic changes (Wharton and others, 1992; Reid & Crout, 2008). Taylor Valley experienced a
370 general cooling trend between 1986 and 1999 (0.7 °C per decade), associated with increasing
371 lake ice thickness (0.11 m a⁻¹) and increasing shortwave radiation (8.1 W m⁻² per decade), based
372 on meteorological data from the Lake Hoare basin (Doran and others, 2002b). Since early- to
373 mid-2000, the increasing ice-thickness trends reversed: WLB ice thickness started to decrease in
374 2002 (Figure 4) whereas the ice thickness of Lake Fryxell started to decline in 2006 (Figure 7).
375 The most notable change in the Taylor Valley climate during this time period was an
376 exceptionally warm summer season during 2001/2002 when maximum air temperature was
377 ~~greater than~~ 9 °C (Doran and others, 2008; Barrett and others, 2008).

378 The influence of seasonal temperature extremes on the long-term ice thickness trends is
379 explored below. A sensitivity experiment was designed where the exceptionally warm season
380 meteorological data of 2001/2002 were substituted with meteorological data from an unusually
381 cold season from the year prior (for WLB). The simulation results show no changes in the ice

382 thickness trend ($r = 0.98$, $p < 0.001$). Conversely, the unusually cold season meteorological data,
383 from a previous year, were randomly forced in the model during the ice thickness decline to test
384 whether it would influence the long-term ice thickness trends. No changes in the trends were
385 detected following this forcing, indicating that the long-term ice thickness trends in Taylor
386 Valley lakes are not particularly responsive to discrete warming/cooling events. Rather, water
387 column heat flux plays a major role in regulating ice thickness of the study lakes.

388 389 Discussion

390 Predictions of long-term ice thickness trends of stratified perennially ice-covered lakes must
391 consider the unique temperature profile of the water column of a lake as an additional heat
392 source. In Antarctica's perennially ice-covered lakes, heat accumulates in the water column due
393 to the continuous influx of solar radiation during austral summers (Chinn, 1993; Shirtcliffe &
394 Benseman, 1964). The deep water temperature maximum associated with higher salt content
395 takes decades to develop (Vincent and others, 2008).

396 Thermal structures of water columns can influence long-term ice thickness trends and
397 preclude a response of the ice thickness to climatic changes. The water column temperature in
398 Taylor Valley lakes is controlled by penetrating solar radiation (Shirtcliffe & Benseman, 1964;
399 Vincent and others, 2008; Spigel & Priscu, 1998). Penetrating solar radiation decays with depth
400 in the ice and the water column based on the extinction coefficients. As a result, shallow water
401 receives more energy than deep water and it will be more responsive to seasonal variations of
402 penetrating solar radiation. Additionally, shallow water temperature is influenced by the
403 contribution of sensible heat flux from the seasonal stream input, which enhances ice thinning.
404 However, it is not included in the model. The energy stored at depth (long-term storage) will
405 affect shallow water temperature, to a certain extent, by the upward heat conduction due to the
406 thermal gradient. As a conceptual example, for a deep lake with a temperature maximum at some
407 depth, during a climate cooling trend (assuming constant shortwave radiation), shallow waters
408 will cool down, increasing ice growth rates. Yet, shallow water will cool at rates slower than that
409 of the air due to an influx of heat from the deep lake heat storage. The incoming heat from
410 beneath will hinder shallow water cooling and slow ice thickness growth (example, Figure 8).
411 Conversely, for shallow lakes that do not have a deep temperature maximum, this process will be
412 less pronounced and the shallow water temperature will more closely reflect climate cooling
413 (Figure 9). The situation would reverse during climate warming; the ice thickness of deep lakes
414 with a well-developed temperature maximum would respond to surface air warming quicker due
415 to the additional heat flux from beneath. For example, the ice thickness of WLB responded
416 earlier to the increase of surface air temperatures (Figures 4 and 8) between 2000 and 2012 in
417 comparison with ice thickness of Lake Fryxell (Figures 7 and 9). This suggests that lakes with
418 deep temperature maxima can reduce the response of ice thickness to climatic changes.

419 The modeled water temperature profile of WLB shows three distinct trends: a continuous
420 increase of deep water temperature over time, a continuous decrease of temperature maximum
421 within about the first 5 meters under the ice cover (Figure 6), and a shallow water temperature
422 (immediately beneath the ice) that varies inversely with the ice thickness ($r = -0.85$, $p < 0.001$)
423 (Figure 8). Annually measured temperature profiles of WLB show similar trends (unpublished
424 data). The only exception in measured data is an increase of temperature maximum after 2009,
425 which is not captured by the model. The cooling of shallow water is associated with an ice
426 thickness increase from 1996 to the end of 2001 (Figure 8). Conversely, the warming of shallow
427 water is associated with an ice thickness decrease after 2001 (Figure 8). A similar trend of

428 inversely proportional ice thickness and shallow water temperature is observed at Lake Fryxell;
429 however, a clear temporal change in ice thickness and water temperature occurred in 2005
430 (Figure 9). The difference in response time of the ice thickness in these two lakes can be
431 attributed to the different thermal structures of the water columns near the ice-water interface,
432 which is influenced by the deep thermal heat storage.

433 The observed long-term ice thickness trends in both lakes are a result, first and foremost, of
434 the thermal structure of the water column and surface energy balance. The infrequent seasonal
435 temperature warming do not influence long-term ice thickness trends because a secondary
436 process (heat flux from the water column) that is driven on much longer timescale influences the
437 long-term trends of ice thickness. The alteration of the ice thickness trend occurs at a threshold
438 dictated by heat fluxes at the ice surface and shallow water temperature, which is influenced by
439 the thermal structure of the water column. Shallow water temperature is controlled by the
440 penetrating shortwave radiation and the heat flux from deep waters. As a result, lake ice
441 thickness will respond to climate forcings on different time scales depending on the thermal
442 structure of the lake in question. We suggest that shallow perennially ice-covered lakes, such as
443 Lake Fryxell, are a better proxy for climate change, owing to the fact that the ice thickness
444 response to long-term climate forcing is not affected by the deep thermal water maximum.

445 446 **Conclusions**

447 A thermodynamic model parameterized for ice cover of WLB was developed. The model is
448 driven by surface radiative heat fluxes and heat fluxes from the underlying water column,
449 coupled with two one-dimensional heat equations. The water temperature profile is calculated in
450 order to properly constrain heat fluxes at the ice/water boundary. Despite the simplistic nature of
451 the model (it does not account for physics associated with isothermal ice, advection of the water
452 column, and sensible heat associated with stream inflow), the model successfully calculates long-
453 term ice thickness changes as well as the temperature profile of the ice and the water column in
454 two stratified lakes. Predicted ice thickness changes are in good agreement with trends observed
455 in measured ice thickness data.

456 Perennially ice-covered lakes with a deep-water temperature maximum will impede the
457 response of ice thickness growth to surface air-cooling. The impact of surface air-cooling is
458 hindered by the heat flux from the water column. Conversely, during surface air warming, the
459 water temperature maximum facilitates ice decay, due to an additional heat flux from below. Ice
460 thickness of shallow lakes (<20 m), on the other hand, will more accurately respond to climatic
461 changes because of diminished effect of the water column temperature on the ice thickness.
462 Hence, ice covers can be used as indicators of climate change; however, the ice thickness trends
463 should be interpreted with caution due to the influence of thermal stratification of the water
464 column.

465 **Acknowledgments**

466 This research was supported by the Office of Polar Programs (grants 9810219, 0096250,
467 0832755, 1041742, and 1115245). Logistical support was provided by the US Antarctic Program
468 through funding from NSF. We thank the Long Term Ecological Research project personnel for
469 collecting the ice thickness measurements throughout the duration of this project and Douglas
470 MacAyeal for insightful conversations on the topic.
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- 472 Adams, E.E., J. Priscu, C.H. Fritsen, S.R. Smith and S.L. Brackman 1998. Permanent ice
473 covers of the McMurdo Dry Valley lakes, Antarctica: bubble formation and metamorphism. In
474 Priscu, J.C., ed. *Ecosystem dynamics in a polar desert; the McMurdo dry valleys, Antarctica.*,
475 American Geophysical Union, 281-295. (Antarctic Research Series.
- 476 Andersen, D.T., C.P. McKay and R.A. Wharton 1998. Dissolved gases in perennially ice-
477 covered lakes of the McMurdo dry valleys, Antarctica. *Antarctic Science*, **10**(2): 124-133.
- 478 Barrett, J.E., R.A. Virginia, D.H. Wall, P.T. Doran, A.G. Fountain, K.A. Welch and W.B. Lyons
479 2008. Persistent effects of a discrete warming event on a polar desert ecosystem. *Global Change*
480 *Biology*, **14**(10): 2249-2261.
- 481 Bennett, T.J. 1982. A Coupled Atmosphere-Sea Ice Model Study of the Role of Sea Ice in
482 Climatic Predictability. *Journal of the Atmospheric Sciences*, **39**(7): 1456-1465.
- 483 Bintanja, R. and M.R. Vandenbroeke 1995. The Surface-Energy Balance of Antarctic Snow and
484 Blue Ice. *Journal of Applied Meteorology*, **34**(4): 902-926.
- 485 Brandt, R.E. and S.G. Warren 1993. Solar-Heating Rates and Temperature Profiles in Antarctic
486 Snow and Ice. *Journal of Glaciology*, **39**(131): 99-110.
- 487 Chinn, T.J. 1993. Physical hydrology of the dry valley lakes. In Green, W.J. and E.I. Friedmann,
488 eds. *Physical and Biogeochemical Processes in Antarctic Lakes*, American Geophysical Union,
489 1-51. (Antarctic Research Series.
- 490 Danard, M., M. Gray and G. Lyv 1984. A Model for Predicting Ice Accretion and Ablation in
491 Water Bodies. *Monthly Weather Review*, **112**(6): 1160-1169.
- 492 Doran, P.T., G.L. Dana, J.T. Hastings and R.A.J. Wharton 1995. McMurdo Dry Valleys Long-
493 Term Ecological Research (LTER): LTER automatic weather network (LAWN). *Antarctic*
494 *Journal of the U.S.*, **30**(5): 276-280.
- 495 Doran, P.T., C.P. McKay, G.D. Clow, G.L. Dana, A. Fountain, T. Nylen and W.B. Lyons 2002a.
496 Valley floor climate observations from the McMurdo Dry Valleys, Antarctica, 1986-2000.
497 *Journal of Geophysical Research*, **107** (D24): 4772, doi:10.1029/2001JD002045.
- 498 Doran, P.T., C.P. McKay, A.G. Fountain, T. Nylen, D.M. McKnight, C. Jaros and J.E. Barrett
499 2008. Hydrologic response to extreme warm and cold summers in the McMurdo Dry Valleys,
500 East Antarctica. *Antarctic Science*, **20**(5): 499-509.
- 501 Doran, P.T., J.C. Priscu, W.B. Lyons, J.E. Walsh, A.G. Fountain, D.M. McKnight, D.L.
502 Moorhead, R.A. Virginia, D.H. Wall, G.D. Clow, C.H. Fritsen, C.P. McKay and A.N. Parsons
503 2002b. Antarctic climate cooling and terrestrial ecosystem response. *Nature*, **415**(6871): 517-
504 520.
- 505 Doran, P.T., R.A. Wharton, Jr. and W.B. Lyons 1994. Paleolimnology of the McMurdo Dry
506 Valleys, Antarctica. *Journal of Paleolimnology*, **10**(2): 85-114.

- 507 Dugan, H.D., M.K. Obryk and P.T. Doran 2013. Lake ice ablation rates from permanently ice-
508 covered Antarctic lakes. *Journal of Glaciology*, **59**(215): 491-498.
- 509 Fountain, A., W.B. Lyons, M.B. Burkins, G.L. Dana, P.T. Doran, K.J. Lewis, D. McKnight, D.L.
510 Moorhead, A.N. Parsons, J.C. Priscu, D.H. Wall, R.A.J. Wharton and R.A. Virginia 1999.
511 Physical controls on the Taylor Valley Ecosystem, Antarctica. *Bioscience*, **49**(12): 961-972.
- 512 Fritsen, C.H. and J.C. Priscu 1999. Seasonal change in the optical properties of the permanent ice
513 cover on Lake Bonney, Antarctica: consequences for lake productivity and phytoplankton
514 dynamics. *Limnology and Oceanography*, **44**(2): 447-454.
- 515 Guest, P.S. 1998. Surface longwave radiation conditions in the eastern Weddell Sea during
516 winter. *Journal of Geophysical Research-Oceans*, **103**(C13): 30761-30771.
- 517 Gupta, S.C. 2003. *The classical Stefan problem : basic concepts, modelling, and analysis*.
518 Amsterdam ; Boston, Elsevier.
- 519 Hoffmann, F.O. and R.H. Gardner 1983. Evaluation of uncertainties in environmental
520 radiological assessment models. US Nuclear Regulatory Commission NUREG/CR-3332.
- 521 Konig-Langlo, G. and E. Augstein 1994. Parameterization of the downward long-wave radiation
522 at the Earth's surface in polar regions. *Meteorol. Zeitschrift*, **3**: 343-347.
- 523 Launiainen, J. and B. Cheng 1998. Modelling of ice thermodynamics in natural water bodies.
524 *Cold Regions Science and Technology*, **27**(3): 153-178.
- 525 Liston, G.E., J. Winther, O. Brunland, H. Elvehoy and K. Sand 1999. Below-surface ice melt on
526 the coastal Antarctica ice sheet. *Journal of Glaciology*, **45**(150): 273-284.
- 527 Lumb, F.E. 1964. The influence of cloud on hourly amounts of total radiation on the sea surface.
528 *Quart. J. R. Met. Soc.*, **90**: 43-56.
- 529 Maykut, G.A. and N. Untersteiner 1971. Some Results from a Time-Dependent Thermodynamic
530 Model of Sea Ice. *Journal of Geophysical Research*, **76**(6): 1550-&.
- 531 McKay, C.P., G.D. Clow, D.T. Andersen and R.A. Wharton 1994. Light transmission and
532 reflection in perennially ice-covered Lake Hoare, Antarctica. *Journal of Geophysical Research*,
533 **99**(C10): 20427-20444.
- 534 McKay, C.P., G.D. Clow, R.A. Wharton and S.W. Squyres 1985. Thickness of ice on perennially
535 frozen lakes. *Nature*, **313**(6003): 561-562.
- 536 Parkinson, C.L. and W.M. Washington 1979. Large-Scale Numerical-Model of Sea Ice. *Journal*
537 *of Geophysical Research-Oceans and Atmospheres*, **84**(Nc1): 311-337.
- 538 Priscu, J.C., M.T. Downes and C.P. McKay 1996. Extreme supersaturation of nitrous oxide in a
539 poorly ventilated Antarctic lake. *Limnology and Oceanography*, **41**(7): 1544-1551.

- 540 Reid, T. and N. Crout 2008. A thermodynamic model of freshwater Antarctic lake ice.
541 *Ecological Modelling*, **210**(3): 231-241.
- 542 Scott, R.F. 1905. *The Voyage of Discovery*. London, McMillan and Co.
- 543 Shirtcliffe, T.G.L. and R.F. Benseman 1964. A sun-heated Antarctic lake. *J. Geophys. Res.*, **69**
544 3355-3359.
- 545 Spigel, R.H. and J.C. Priscu 1996. Evolution of temperature and salt structure of Lake Bonney, a
546 chemically stratified Antarctic lake. *Hydrobiologia*, **321**(3): 177-190.
- 547 Spigel, R.H. and J.C. Priscu 1998. Physical limnology of the McMurdo Dry Valley lakes. In
548 Priscu, J.C., ed. *Ecosystem Dynamics in a Polar Desert: The McMurdo Dry Valleys, Antarctica*,
549 Washington, D.C., American Geophysical Union, 153-187.
- 550 Trefethen, L.N. 2000. *Spectral methods in MATLAB*. Philadelphia, Pa., Society for Industrial and
551 Applied Mathematics.
- 552 Vincent, A.C., D.R. Mueller and W.F. Vincent 2008. Simulated heat storage in a perennially ice-
553 covered high Arctic lake: Sensitivity to climate change. *Journal of Geophysical Research*,
554 **113**(C4).
- 555 Wharton, R.A., C.P. McKay, G.D. Clow, D.T. Andersen, G.M. Simmons and F.G. Love 1992.
556 Changes in Ice Cover Thickness and Lake Level of Lake Hoare, Antarctica - Implications for
557 Local Climatic-Change. *Journal of Geophysical Research*, **97**(C3): 3503-3513.
- 558 Wharton, R.A., C.P. McKay, G.M. Simmons and B.C. Parker 1986. Oxygen Budget of a
559 Perennially Ice-Covered Antarctic Lake. *Limnology and Oceanography*, **31**(2): 437-443.
- 560 Wharton, R.A., G.M. Simmons and C.P. McKay 1989. Perennially Ice-Covered Lake Hoare,
561 Antarctica - Physical-Environment, Biology and Sedimentation. *Hydrobiologia*, **172**: 305-320.
- 562 Willmott, C.J. 1981. On the validation of models. *Physical Geography*, **2**(2): 184-194.
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564

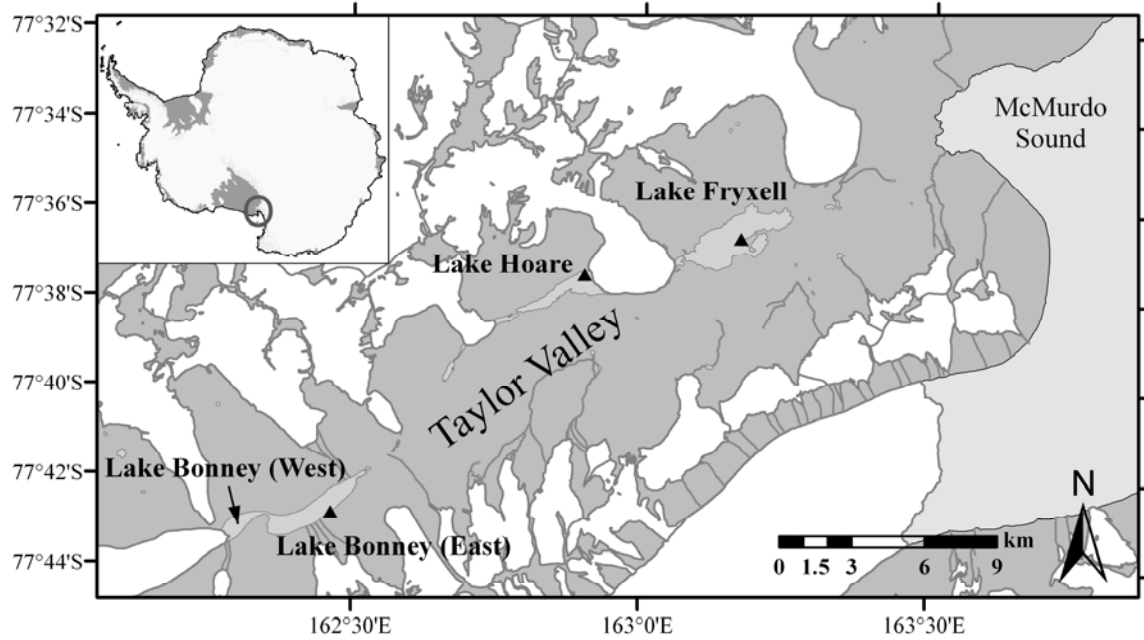


Figure 1. Map of Taylor Valley, Antarctica. Triangles indicate locations of meteorological stations.

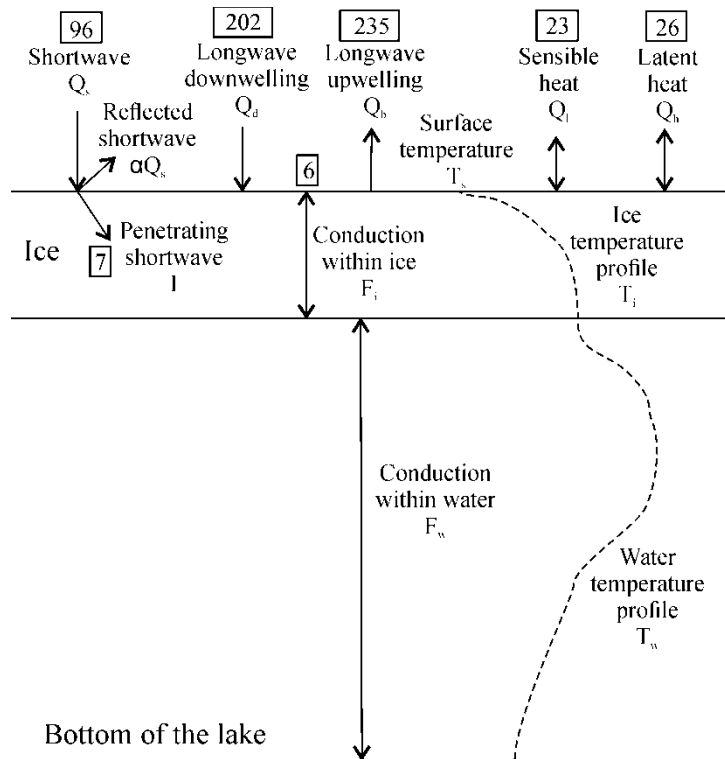


Figure 2. Conceptual representation of heat fluxes and temperature profile for west lobe of Lake Bonney. Numbers in boxes represent annually averaged fluxes (W m^{-2}).

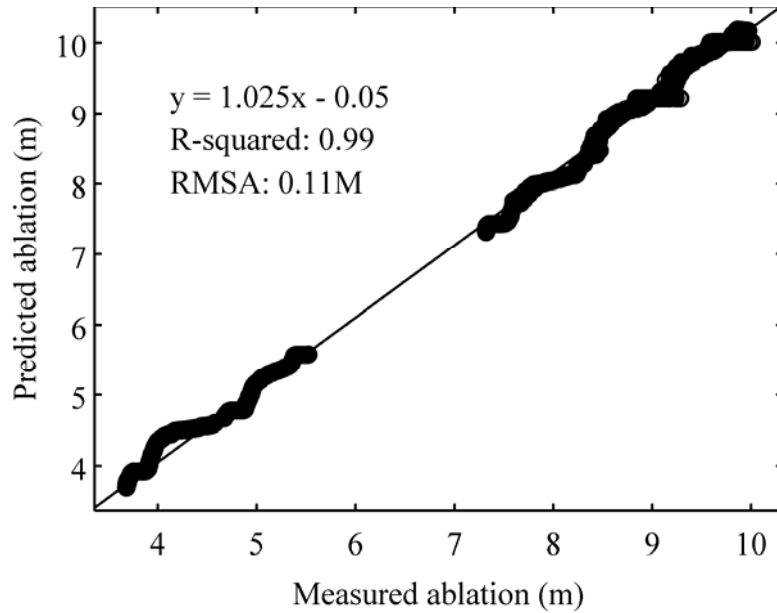


Figure 3. Linear fit between 6 years of overlapping measured and predicted ice ablation for west lobe of Lake Bonney (each data point represent 12-hr averages, $n = 4251$). ~~RMSE is Root Mean Square Error.~~

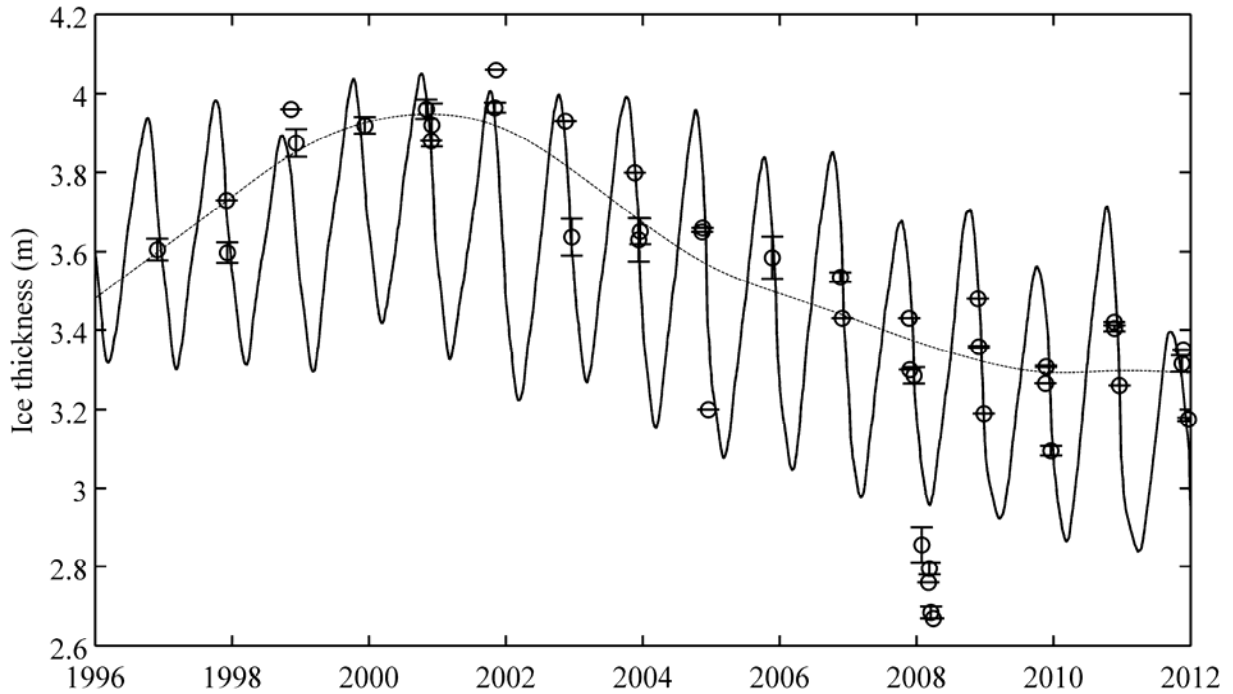


Figure 4. Predicted ice thickness changes for west lobe of Lake Bonney (black solid line) and averaged measured ice thickness (circles) from 1996 to 2012. Numerous ice thickness measurements were obtained during the same day, showing large variability. Days with multiple ice thickness measurements were averaged. Error bars are shown as standard deviation. Dashed line represents a spline fit to measured data, excluding late season 2008 measurements.

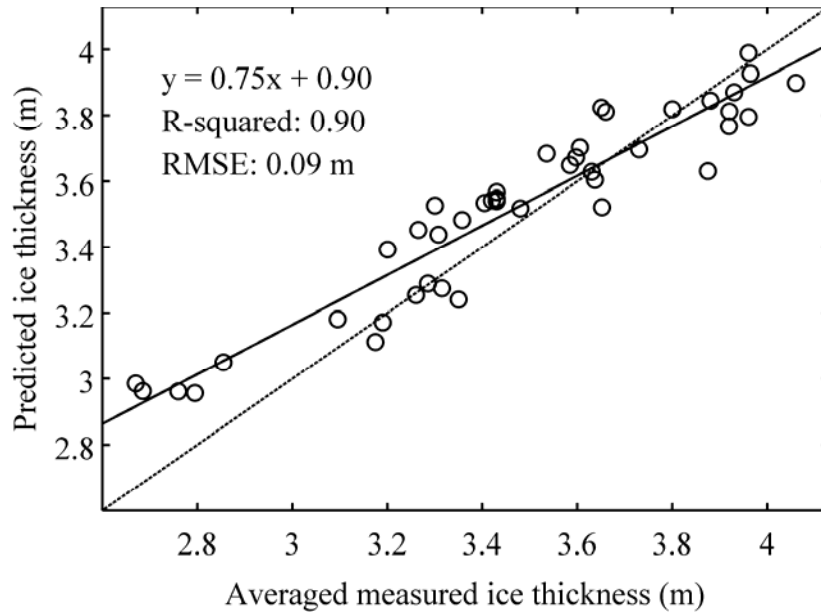


Figure 5. Linear fit between 16 years of overlapping measured and predicted ice thickness changes for west lobe of Lake Bonney. Circles represent daily averaged overlapping data. ~~RMSE~~ is ~~Root-Mean-Square-Error~~. Dashed line is 1:1.

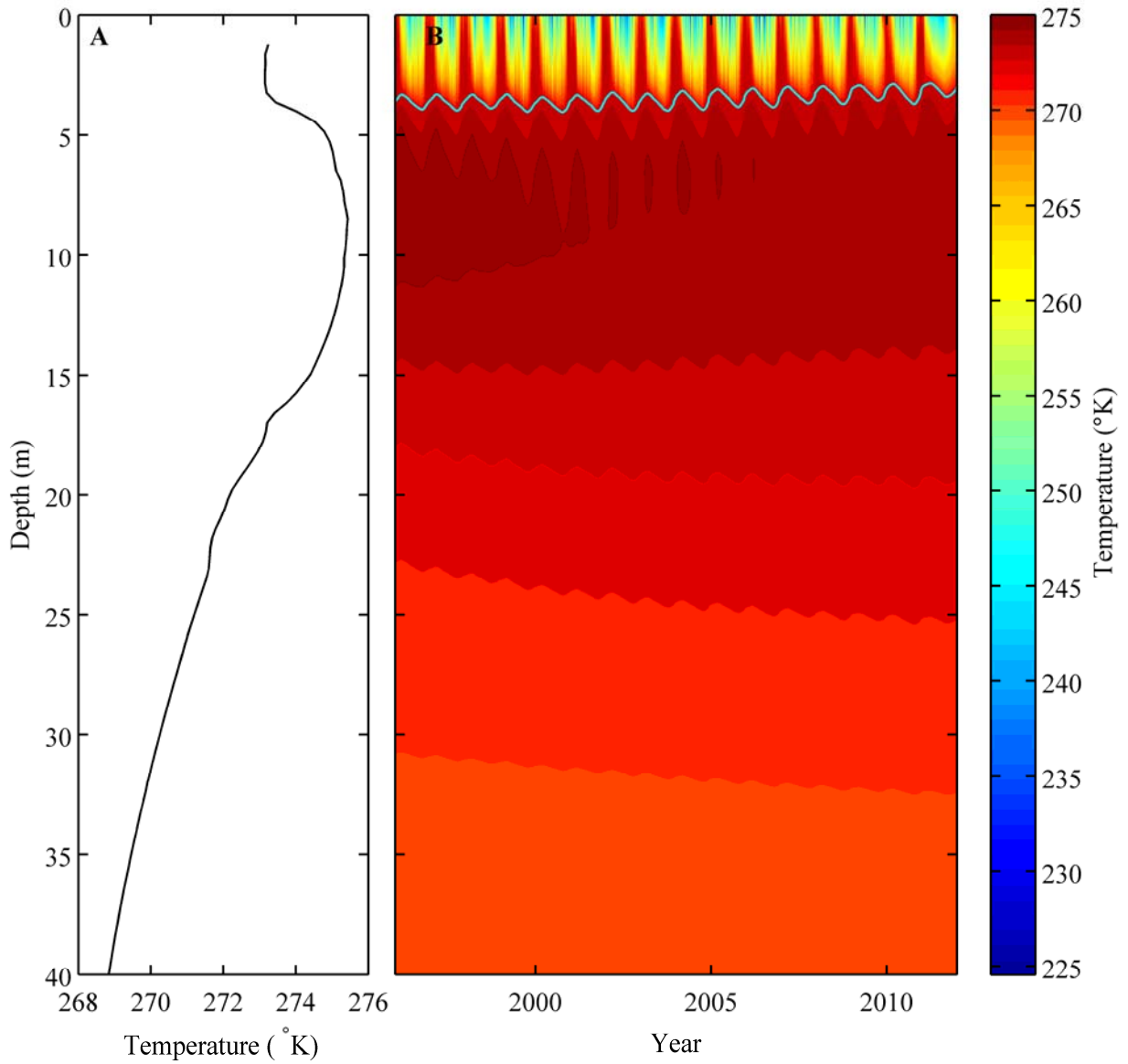


Figure 6. **A)** A typical water temperature profile from west lobe of Lake Bonney (data obtained on Nov 21st 2011). **B)** Modeled evolution of water temperature profile between 1996 and 2012.

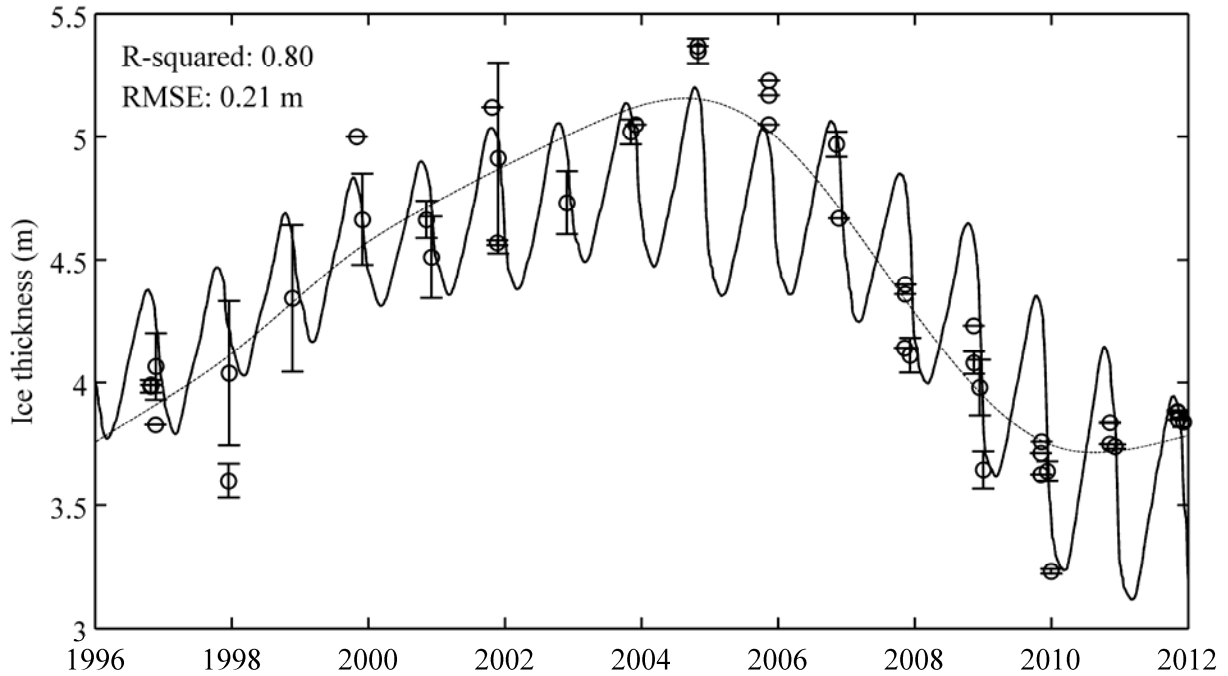


Figure 7. Predicted ice thickness changes for Lake Fryxell (solid line) and averaged measured ice thickness (circles) from 1996 to 2012. Numerous ice thickness measurements were obtained during the same day, showing large variability. Days with multiple ice thickness measurements were averaged. Error bars are shown as standard deviation. ~~RMSE - Root-Mean-Square-Error.~~ Dashed line represents a spline fit to measured data, excluding late season 2008 measurements.

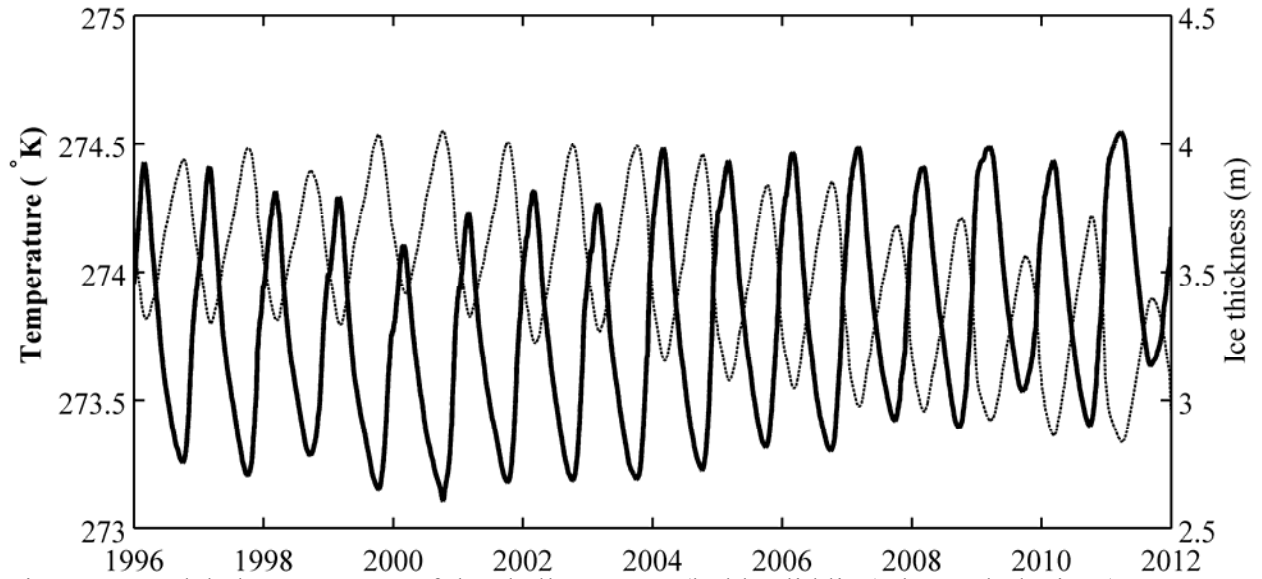


Figure 8. Modeled temperature of the shallow water (bold solid line), beneath the ice (4 m depth), over time at west lobe of Lake Bonney and modeled ice thickness change over time (dashed line). Shallow water temperature and ice thickness changes are inversely proportional.

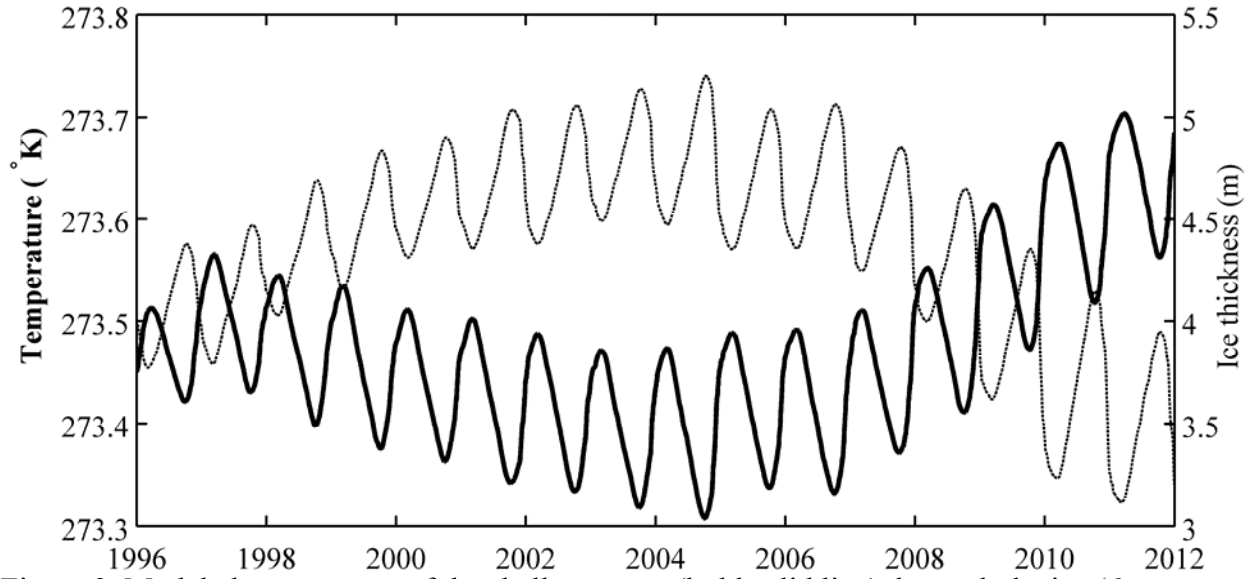


Figure 9. Modeled temperature of the shallow water (bold solid line), beneath the ice (6 m depth), over time at Lake Fryxell and modeled ice thickness change over time (dashed line). Shallow water temperature and ice thickness changes are inversely proportional.

Table 1. Sensitivity Index (Si) results for parameters used in the model. Values less than 0.01 represent no sensitivity to the changes, whereas Si of 1 indicates large sensitivity to the changes. χ wavelength depended absorption of shortwave radiation at the surface of the ice, α albedo, Q_s shortwave radiation, T_a air temperature, C_E bulk transfer coefficient used in latent heat equation, C_H bulk transfer coefficient used in the sensible heat equation, κ extinction coefficient.

	χ	α	Q_s	T_a	C_E	C_H	κ
Si	0.46	0.33	0.27	0.19	0.06	< 0.01	< 0.01